Earth Surface Processes and Landforms
Earth Surf. Process. Landforms, **23**, 825–843 (1998)

SEASONAL CHANGES IN THE MORPHOLOGY OF THE SUBGLACIAL DRAINAGE SYSTEM, HAUT GLACIER D'AROLLA, SWITZERLAND

PETER NIENOW[1], MARTIN SHARP[2] AND IAN WILLIS[3]
[1]*Department of Geography and Topographic Science, University of Glasgow, Glasgow, G12 8QQ, UK*
[2]*Department of Earth and Atmospheric Sciences, University of Alberta, Edmonton, Alberta T6G 2E3, Canada*
[3]*Department of Geography, University of Cambridge, Downing Place, Cambridge, CB2 3EN, UK*

Received 6 July 1994; Revised 21 January 1998; Accepted 21 January 1998

ABSTRACT

Dye tracing techniques were used to investigate the glacier-wide pattern of change in the englacial/subglacial drainage system of Haut Glacier d'Arolla during the ablation seasons of 1990 and 1991. Analysis of breakthrough curve characteristics indicate that over the course of a melt season, a system of major channels developed by headward growth at the expense of a hydraulically inefficient distributed system. By the end of the melt season, this channel system extended at least 3·3 km from the snout of the 4 km long glacier and drained the bulk of supraglacially derived meltwater passing through the glacier. The upper limit of the channel system closely followed the retreating snowline up-glacier. Rates of headward channel growth reached *c*. 65 m d^{-1}, although these rates decreased in the upper 1 km of the glacier where snowline retreat exposed a patchy firn aquifer. It appears that the removal of snow (with its high albedo and significant water storage capacity) from the glacier surface resulted in a dramatic increase in the volume of runoff into moulins, and in the peakedness of daily runoff cycles. This induced transient high water pressures within the distributed drainage system, which caused it to evolve rapidly into a channelised system. It is therefore likely that, at a local scale, channel growth occurred down-glacier from moulins, and that the overall up-glacier-directed pattern of channel formation was caused by the retreating snowline exposing new moulins and crevasses to inputs of ice-derived meltwater. Damping of diurnal melt inputs by storage in the firn aquifer accounts for the slowing of channel growth in the upper glacier. © 1998 John Wiley & Sons, Ltd.

KEY WORDS: glacier hydrology; subglacial drainage systems; dye tracing

INTRODUCTION

Periodic glacier flow instabilities such as surges may be linked to temporal changes in the morphology of subglacial drainage systems (Kamb *et al*., 1985; Kamb, 1987). This suggestion has aroused considerable interest in the manner in which such drainage systems evolve on shorter (seasonal and diurnal) time scales, and in the processes by which this evolution takes place (e.g. Hock and Hooke, 1993; Stone and Clarke, 1996). These phenomena are also important for understanding water storage in and runoff from glaciers. Since the storage capacities of different types of drainage system may vary considerably, processes of drainage evolution may play an important role in tapping subglacially stored water (Willis *et al*., 1993). Since the majority of bulk runoff from temperate glaciers is derived from meltwater generated at the glacier surface, it is likely that changes within the englacial/subglacial drainage system are linked directly to surface processes that control how meltwater is delivered to it. Thus seasonal changes in the shape of bulk runoff hydrographs from glaciers may reflect both the effect of melt removal of the supraglacial snowpack on water storage and routing to the englacial/subglacial system (Fountain, 1996) and the consequent changes in the efficiency and morphology of the system (Röthlisberger and Lang, 1987).

In an attempt to determine whether, how and why an englacial/subglacial drainage system changes on the scale of a whole glacier over a summer melt season, 415 dye tracer experiments were conducted at Haut Glacier

* Correspondence to: P. Nienow
Contract/grant sponsor: NERC; grant number: GR3/7004a.
Contract/grant sponsor: Earthwatch

CCC 0197-9337/98/090825–19 $17.50

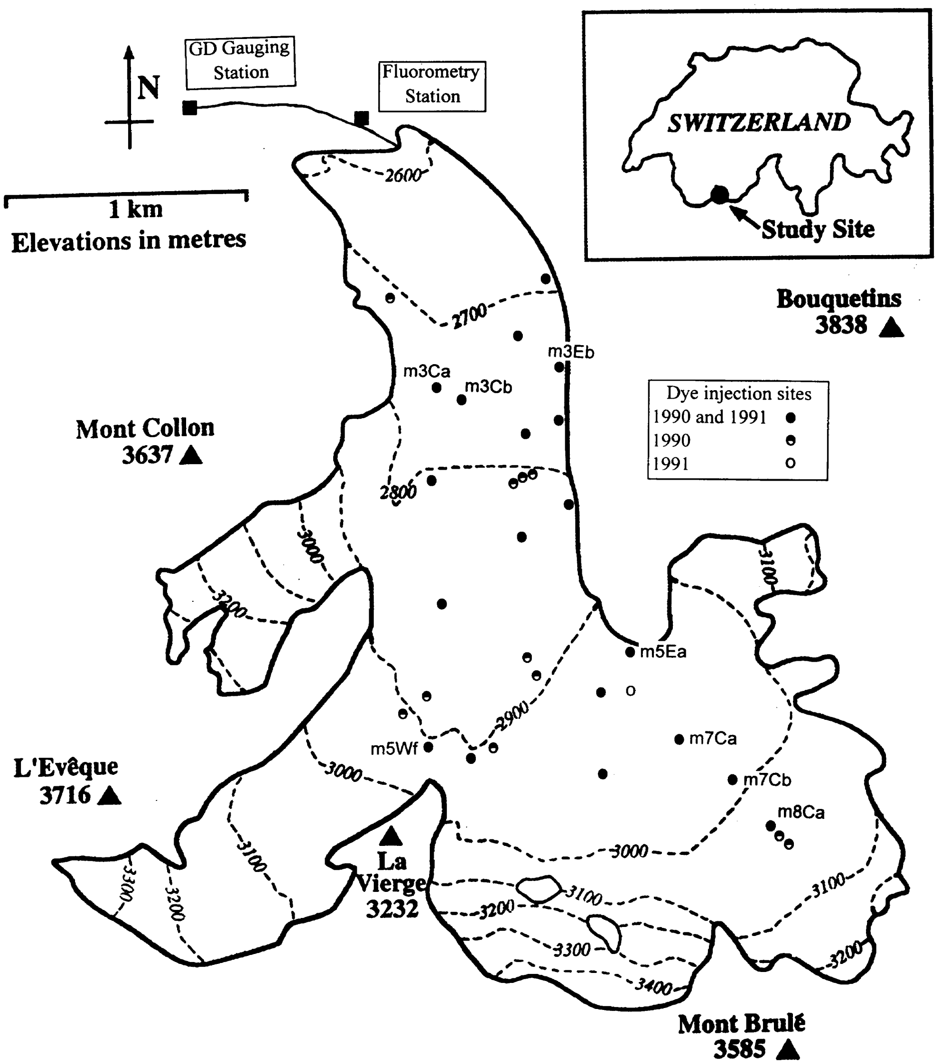

Figure 1. Map of Haut Glacier d'Arolla, showing the location of dye injection sites used during the summers of 1990 and 1991, the fluorometry station and the Grand Dixence gauging station. The labelled moulins are referred to in the text. A four-digit code is used to classify sites as either moulins (m) or extraglacial streams (s), to show their longitudinal (1–8 in 500 m long segments, with 1 being the segment closest to the glacier snout) and transverse (E, C. W) position on the glacier, and their relative proximity to the glacier snout in a given segment (a to z, with a being closest to the snout)

d'Arolla, Switzerland (Figure 1), during the summers of 1990 and 1991. In both years, the tracer injection programme began as soon as runoff was detected at the glacier surface and continued until late August (1990) or early September (1991).

METHODS

Dye injection and detection

Efforts were made to initiate the dye injection programme at moulins which lay significantly above the transient snowline, and to make repeat injections from each site at regular intervals through the melt season. Injection sites were distributed throughout the 4 km length of the glacier, from within 500 m of the glacier snout to within 400 m of the glacier headwall (Figure 1). Rhodamine-B was used as a tracer in both seasons and

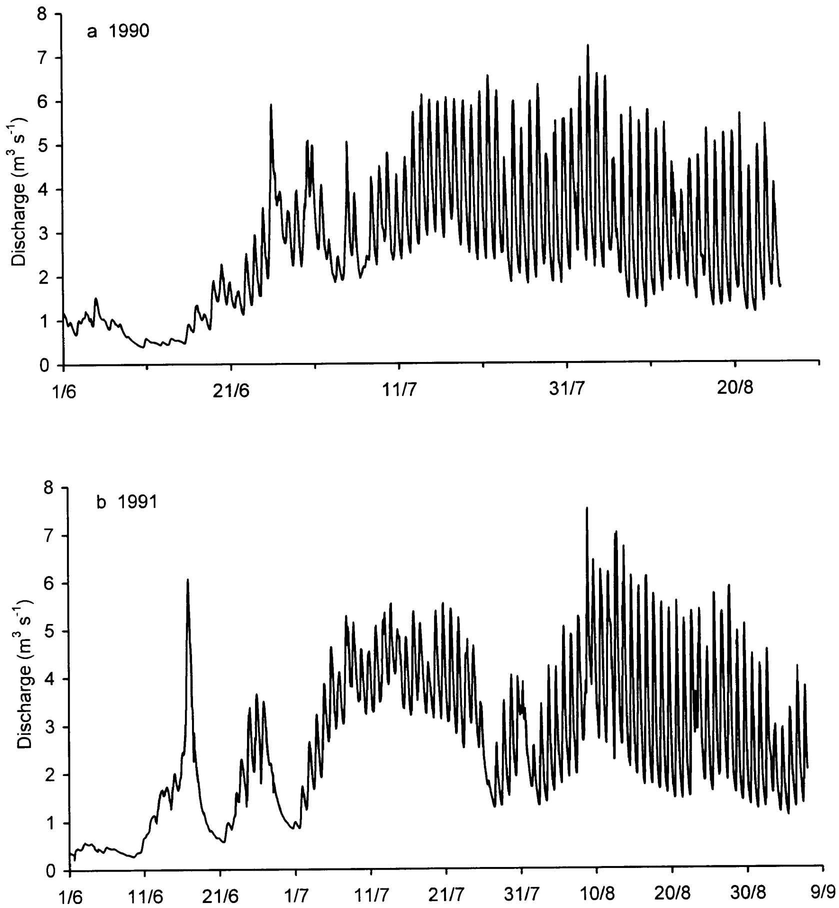

Figure 2. Meltwater discharge records for the Haut Glacier d'Arolla for the summers of (a) 1990 and (b) 1991. Data provided by Grande Dixence SA

fluorescein was also used in 1990. Dye injection involved flushing a known quantity of dye in solution into a freely draining moulin or crevasse. When necessary, snow was excavated from a moulin or crevasse so that dye was injected directly into the englacial drainage system without delay occurring in the snowpack. Dye emergence was detected by continuous flow fluorometry (Sharp *et al*., 1993) the location shown on Figure 1.

Discharge measurement

Measurements of bulk discharge (Q_b) were obtained from the Grande Dixence SA hydroelectric intake structure located 950 m downstream from the glacier snout (Figures 1 and 2). The accuracy of these measurements is ±4% (Brown and Tranter, 1990). The local meltwater discharge (Q_l) through the glacier cross-section at each injection site was estimated using a knowledge of the glacier hypsometry and observed patterns of variation in melt rates with surface elevation (Nienow, 1993). Mean discharge (Q_m) between the injection site and the glacier snout was taken to be $Q_m = [(Q_{b1} + Q_{b2}/2) + (Q_{l1} + Q_{l2}/2)]/2$, where the subscripts 1 and 2 refer to discharges at the beginning and end of the tracer test respectively. Since Q_l is a constant fraction of Q_b, it follows that temporal variations in Q_m follow those in Q_b. The rationale for calculating Q_m is that by using this parameter, rather than Q_b, to calculate the cross-sectional area of the drainage system (see below), one is less likely to overestimate the cross-sectional area.

Analysis of dye breakthrough curves

Five main parameters, which provide a measure of the efficiency and morphology of the englacial/subglacial drainage system, were derived from the dye breakthrough curves:

(i) the return time (t) between dye injection and peak dye concentration at the detection site;
(ii) a minimum estimate of the mean flow velocity (u) during the test (x/t, where x is the straight-line distance between injection and detection sites);
(iii) the dispersion coefficient (D, $m^2\ s^{-1}$), which describes the rate of dispersion of the dye cloud during its passage through the glacier (this was calculated according to the method of Seaberg *et al* (1988, equation 4, p. 222);
(iv) the dispersivity ($d = D/u$), which describes the rate of dispersion of the dye cloud relative to its rate of advection through the glacier, and which has been used to infer the complexity of the drainage path followed by the dye (e.g. Seaberg *et al*., 1988; Hock and Hooke, 1993; Hooke and Pohjola, 1994); and
(v) the apparent mean cross-sectional area ($A_m = Q_m/u$) of the englacial/subglacial system between the injection site and the glacier snout. Given the assumptions which underlie the estimates of Q_m and u, A_m provides an index, rather than a true measure, of cross-sectional area.

Due to problems of background fluorescence resulting from suspended sediment in the proglacial stream, dye breakthrough curves obtained from fluorescein experiments could not be used to derive dispersion coefficients (and thus dispersivity).

RESULTS

Individual injection sites

Since experiments from sites initially located above the snowline began relatively earlier in the melt season than those from sites at which injections began after removal of the snowpack, this analysis focuses on results from the above snowline tests. In an attempt to reduce the influence of diurnal discharge variations on the results from tracer tests (Nienow *et al*., 1996a), only tests conducted during the period of peak daily discharge (12.00–17.00) are considered here. In addition, only tests conducted in the main western catchment, which drains approximately 80 per cent of the bulk discharge (Sharp *et al*., 1993), are considered.

Breakthrough curve characteristics

Figure 3a shows the relationship between return time and date of injection for sites which were initially located above the snowline in 1990 and 1991. Return times from each site decreased over time before reaching a relatively stable value. Stabilisation of return times occurred later in the season with increasing distance up-glacier. Flow velocities from 1990 injection sites located up to 3·3 km from the snout therefore increased rapidly over time before stabilising at values in the range 0·37–0·72 $m\ s^{-1}$ (Figure 3b). It is possible that, in 1990, the return times and velocities from the two most distant sites (moulins m7Cb and m8Ca) never reached a stable value (Figure 3a). In 1991, flow velocities from all injection sites increased during the summer, reaching maximum values of between 0·45 and 0·55 $m\ s^{-1}$ (Figure 3b).

These changes in return time and flow velocity were associated with changes in the shape of the dye breakthrough curves. Breakthrough curves from injections made at m3Ca between 17 June and 4 July 1990 became less dispersed and increasingly concentrated into a single peak over time (Figure 4) whilst return time decreased and flow velocity increased (Figure 5a and b). The breakthrough curves from injections on 17 and 19 June showed peaks/shoulders on both rising and falling limbs. The curve from 20 June had a shoulder on its falling limb, whilst the injection on 4 July generated a single peak. Over this period, the dispersion coefficient initially rose from around 7 $m^2\ s^{-1}$ to between 12 and 13 $m^2\ s^{-1}$, before falling to around 3 $m^2\ s^{-1}$ (Figure 5c), while the dispersivity fell from 71·0 m on 17 and 19 June to only 5·8 m on 4 July (Figure 5d). Thus the initial fall in dispersivity was driven primarily by the increase in flow velocity, while the final stages of the decrease resulted from both increasing flow velocity and decreasing dispersion coefficient. Dispersivities calculated from breakthrough curves from the five additional 1990 injection sites showed similar declines through the melt

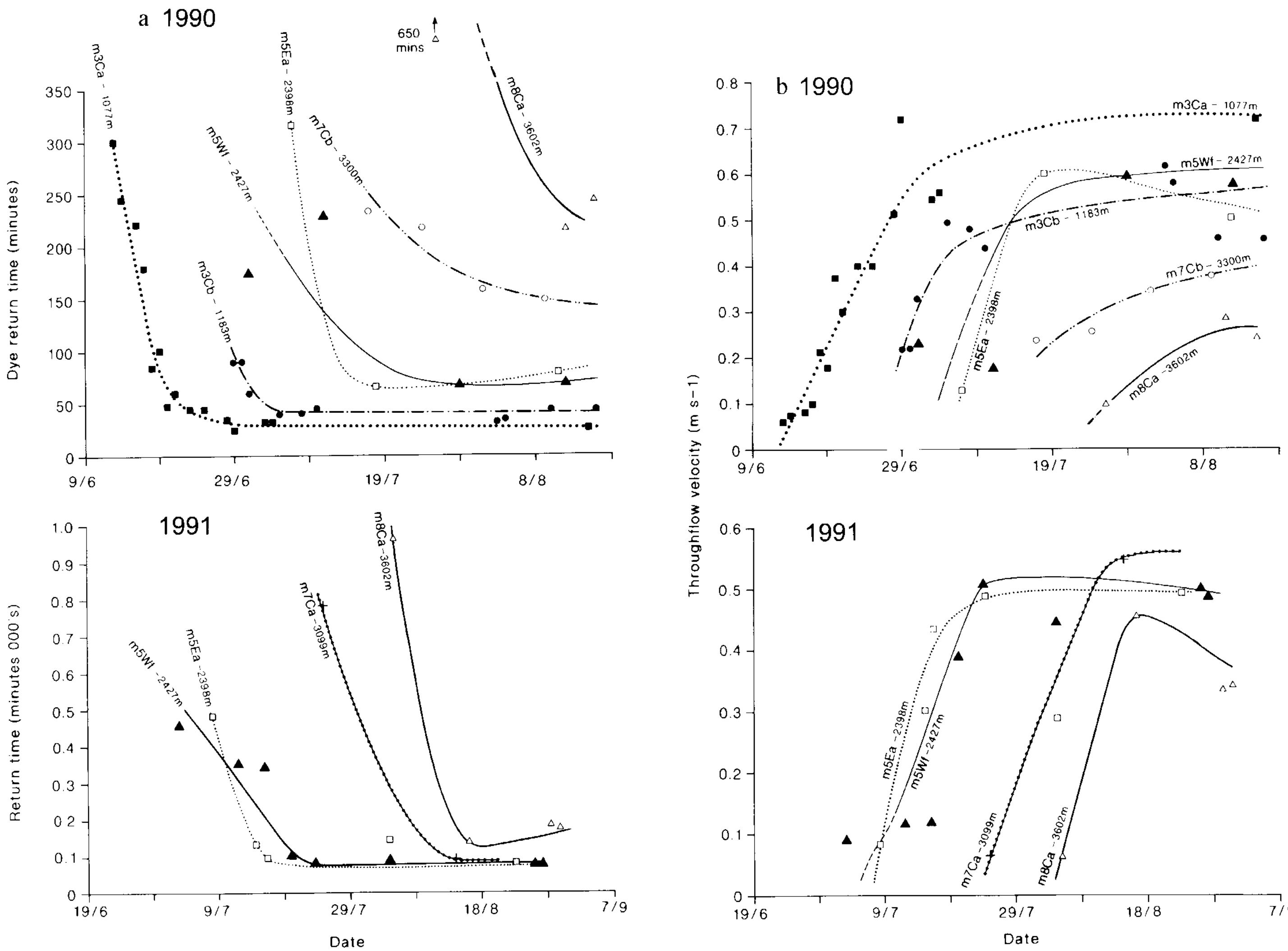

Figure 3. Plots of (a) dye return time and (b) flow velocity as a function of date of injection for selected moulins during the summers of 1990 and 1991. For days on which more than one injection was made at a given site, the shortest return time/highest velocity is plotted. Lines added by hand to aid interpretation

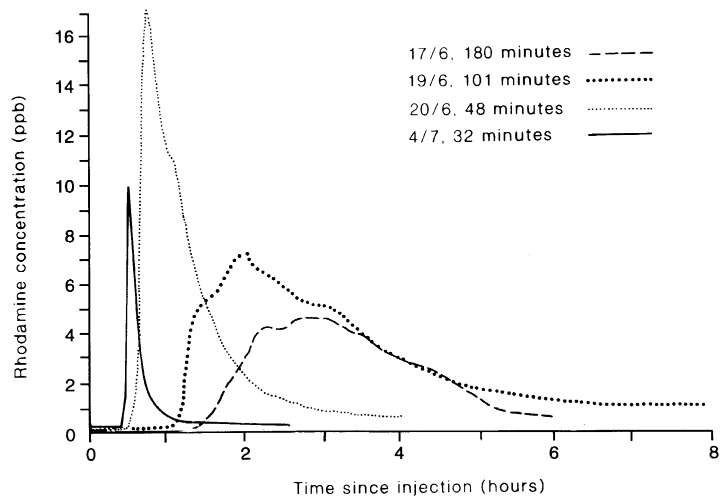

Figure 4. Sequence of four return curves derived from injections made at moulin m3Ca between 17 June and 4 July 1990

season (Figure 6). As with return time and velocity, the major change in dispersivity occurred later in the summer at injection sites located further up-glacier (Figure 6). For all but two of the injection sites, the dispersivity declined to values between 2·8 and 7·5 m. For moulins m7Cb and m8Ca, however, the final recorded values of dispersivity were 34·3 m and 15·8 m on 9 and 15 August respectively.

Cross-sectional area

Continuity arguments suggest that the cross-sectional area of a drainage system (A) can be determined from the ratio between the water discharge (Q) and the flow velocity through the system (u). Problems arise in applying this approach to determine values of A in englacial/subglacial drainage systems because (i) flow velocities determined by dye tracing methods are straight-line averages over the whole flow path and may differ significantly from local flow velocities at particular points in the system, and (ii) measured discharge is usually taken as either the total discharge draining the glacier, or the discharge from an individual distributary channel. Use of these values to calculate A will normally result in overestimation for sites up-glacier from the glacier snout, where discharge is actually lower than assumed (Hock and Hooke, 1993). To avoid the latter problem, we have estimated the local discharge (Q_1) as outlined earlier, and used estimates of mean local discharge during each dye trace (Q_{1m}) to determine the mean cross-sectional area (A_{lm}) at an up-glacier distance equal to that of the injection site. We see no alternative but to use the average velocity determined from tracer tests in these calculations, but recognise that this approach may well overestimate the local velocity at the injection site. If this is the case, our estimate of A_{lm} will be too low. We also emphasise that if drainage at any given distance from the snout occurs via multiple paths (as suggested by the multiple-peaked dye breakthrough curves), the cross-sectional area of individual channels will be less than the total cross-sectional area of the drainage system. Despite these problems, however, there was a clear pattern of change in A_{lm}, which decreased over time at all the injection sites before stabilizing (Figure 7).

Discussion

These results suggest that over the course of the melt season the englacial/subglacial drainage system below each injection site changed from a configuration characterised by low flow velocities, multiple links and large total cross-sectional area to one characterised by higher velocities and fewer links with smaller total cross-sectional area. Rationalisation of the drainage network is implied by the decrease in dispersivity which accompanied these changes, since the dispersivity provides a measure of the characteristic length scale of the drainage system (Fischer, 1968). These changes occurred during a period when total runoff from the glacier was increasing (Figure 2). Under these circumstances, for the flow velocity to increase while the total cross-

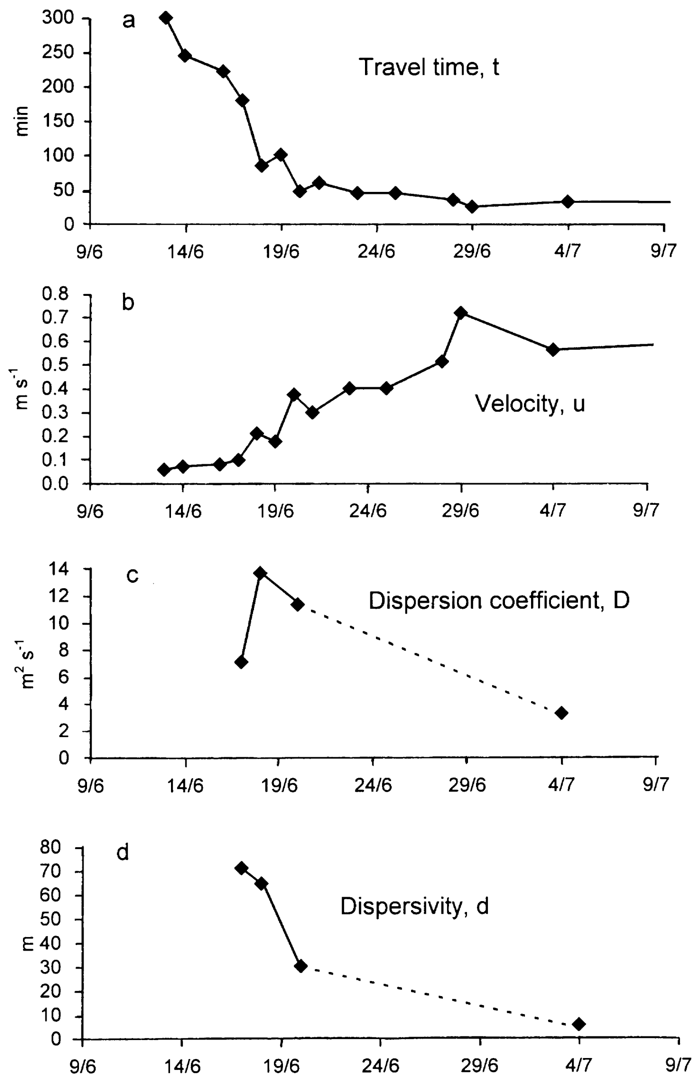

Figure 5. Plots of (a) dye return time, (b) flow velocity, (c) dispersion coefficient, and (d) dispersivity as a function of date of injection for moulin m3Ca during summer 1990. The poor record of dispersion coefficient results from problems involved in analysing breakthrough curves derived from fluorescein tracer tests (see Methods)

sectional area of the drainage system decreased, the hydraulic efficiency of the drainage system must increase. Such an increase in hydraulic efficiency might result from an increase in the hydraulic gradient driving the flow, or from a decrease in the sinuosity or hydraulic resistance of the drainage system. Increased hydraulic gradient could result from either an increase in water pressure within the drainage system or a decrease in sinuosity, as might be implied by the observed changes in dispersivity. Decreased hydraulic resistance could arise from changes in link size, shape or boundary roughness. Given the evidence for decreased total cross-sectional area, and for a reduction in the number of links in the drainage system, it seems likely that individual links within the rationalised drainage system were larger, though fewer in number, than individual links within the system which existed at the start of the measurement period. Velocities within the more efficient system were

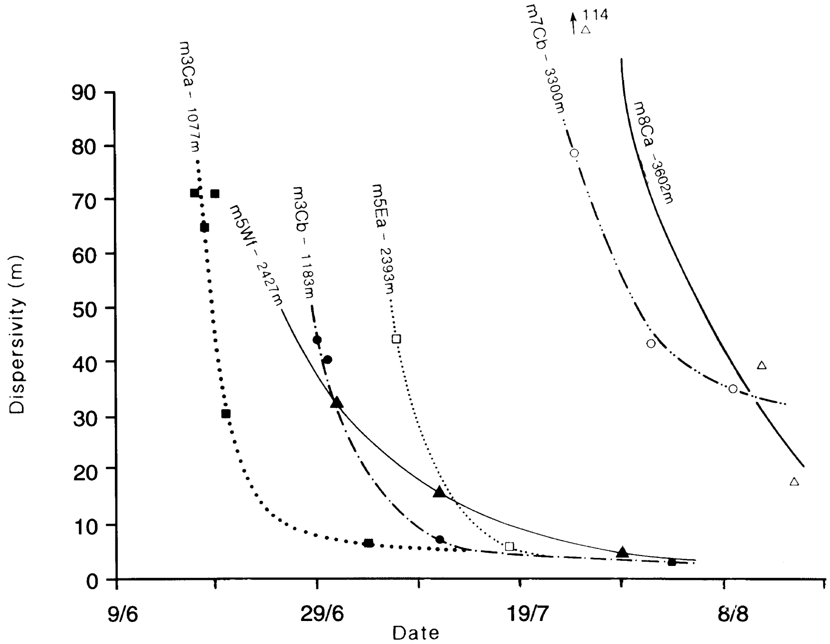

Figure 6. Plots of dispersivity as a function of date of injection for selected moulins during the summer of 1990. Lines added by hand to aid interpretation

sufficiently high to suggest that it consisted of relatively large channels (Behrens *et al*., 1975; Burkimsher, 1983). It thus seems reasonable to suggest that the drainage configuration evolved from a distributed, multi-thread configuration at the start of the melt season, to a more efficient system containing fewer threads by the end of the season.

The change in drainage configuration took place later in the summer at locations further from the glacier snout. This implies that, over the course of the melt season, the area drained by the channelised system increased, while that drained by the distributed system decreased. Once headward growth of the channelised system had begun, water entering the glacier at locations up-glacier from its head probably followed a composite flow path through the glacier. This would involve an initial period of drainage through the distributed system, followed by a period of flow through the channelised system. The decrease in travel time and associated increase in flow velocity observed at each injection site may thus indicate an increase in the proportion of the flow path from that site which lay within the channelised system.

The increase in dispersion coefficient which occurred as flow velocity initially increased (Figure 5b and c) may also be a result of the development of a composite flow path. Since there is an order of magnitude increase in flow velocity between the distributed and channelised systems, a dye cloud passing from the distributed to the channelised system would be rapidly dispersed as its leading edge was advected rapidly through the channelised system, while its trailing edge continued to move slowly through the distributed system. This effect would be unimportant if the drainage system were entirely distributed or channelised, when the magnitude of the dispersion coefficient would reflect the range of velocities within individual drainage links and the complexity of the drainage network structure.

These changes in drainage configuration may have occurred within either, or both, the englacial and subglacial components of the drainage system. Since dye was always injected into single streams, multiple peaked dye breakthrough curves require that, early in the melt season, the flow of these streams became divided at some point between the injection site and the glacier snout. Englacially, flow division could occur if moulins were located along crevasses, since these could store both water and dye (Fountain, 1993), allow bi-directional flow along the crevasse from the injection point, and allow the dye to leave the crevasse via multiple outflow points. If, however, the inflow was not divided englacially, it must have been divided at the bed. This would be

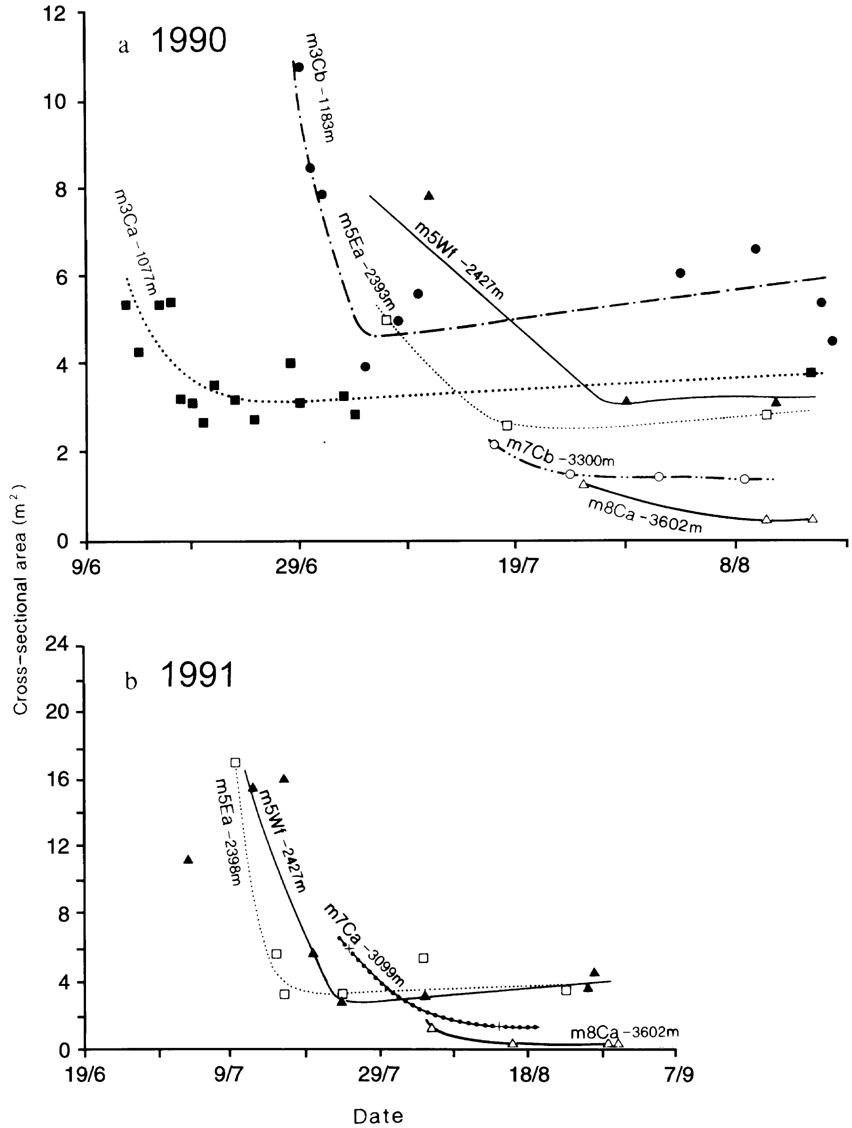

Figure 7. Plots of the reconstructed cross-sectional area of the drainage system at an up-glacier distance equivalent to that of selected moulins as a function of date of injection for the summers of (a) 1990 and (b) 1991. Lines added by hand to aid interpretation

most likely to occur if the vertical dimensions of roughness elements on the bed were large relative to the depth of the flow, such that the flow became divided and diverted. The precise form of this low velocity, multi-thread system is unclear, but possible configurations include a braided sheet flow across a bedrock or sediment surface, or a system of linked cavities in which flow is slowed by storage within cavities.

Rationalisation of the englacial drainage system might result from preferential growth of a single outlet from a crevasse, such that water and dye storage within crevasses was reduced. Rationalisation of the subglacial system also implies preferential development of a small number of drainage links. This could simply involve a reduction in the degree of braiding, as suggested by Hock and Hooke (1993). It might, however, also be facilitated by increased separation between the glacier and its bed resulting from increased sliding and uplift at

times of high subglacial water pressure (Iken, 1981; Iken *et al*., 1983), by melting of the walls of incipient drainage channels, which would be greatest around the largest passageways (Walder, 1982), and by a reduction in the rate of channel closure by ice deformation at times of high water pressure. It could be linked to rationalisation of the englacial drainage system if this resulted in larger fluxes of water being delivered to the glacier bed at a smaller number of points.

Whilst it is impossible to rule out rationalisation of the englacial drainage system, a number of arguments suggest that at least some of the inferred change in drainage configuration must have involved the subglacial system. It is clear from the dye tracing results that the drainage configuration at the start of one melt season is very different from that at the end of the previous one. This implies that shrinkage and closure of drainage passageways takes place over winter. This may result from plugging of crevasses and moulins with snow or refrozen meltwater, or from closure of channels by ice deformation. However, since the stresses driving ice deformation are maximum in the vicinity of the glacier bed, channel closure by this process is most likely to occur at or immediately above the bed. It may be enhanced by the advection by ice flow of abandoned channel segments incised into the base of the glacier into high pressure areas upstream of bedrock obstacles.

A simple calculation suggests that, over most of the glacier, subglacial channels are unlikely to survive from one summer to the next. Following Hooke (1984, equation 7), the closure rate, d*r*/d*t*, of a semi-circular channel at atmospheric pressure can be approximated by:

$$\mathrm{d}r / \mathrm{d}t = (\rho_{\mathrm{i}}\, g h / n_2 A_2)^{n_2}\, r \tag{1}$$

where $n_2 = 3$ and $A_2 = 5{\cdot}8 \times 10^7$ Pa s$^{-1/3}$ (Spring and Hutter, 1982), which are constants from Glen's flow law assuming homogeneous isotropic ice, ρ_i is ice density, *g* is gravitational acceleration, *h* is the ice thickness and *r* is the radius of a semi-circular channel. To obtain an approximate estimate of *r*, we assume that at the end of a melt season, the main catchment is drained by two main channels (Sharp *et al*., 1993) which are semi-circular and each transport 40 per cent of the bulk discharge from the glacier. If we further assume that these channels are adjusted to each transport a mean discharge of 2·4 m^3 s^{-1} flowing at 0·5 m s^{-1} (reasonable late August values for both parameters), the cross-sectional area, *A*, of each channel at the snout would be 4·8 m^2. *r* can be obtained from:

$$r^2 = (2A) / \pi \tag{2}$$

giving a value of 1·75 m. Both *A* and *r* would decrease up-glacier in a manner consistent with the inferred up-glacier decrease in local discharge.

On the assumption that channels close over the 8 months from October to May (due to a lack of runoff and thus wall melting), the degree of channel closure was determined by integration of Equation 1 over time. Calculations were conducted for points located at 100 m intervals along a channel. Ice thicknesses at each point were determined from radio-echo sounding data (Sharp *et al*., 1993). These calculations suggest that channels would survive the winter in the lower 700 m of glacier, where ice thicknesses are less than approximately 90 m. By late May, these channels would have cross-sectional areas of about 0·5 m^2 at a point 700 m from the glacier snout and 4·2 m^2 100 m from the snout. Closure rates in the lowermost 500 m of the glacier would be minimal because ice thicknesses there are less than 50 m. If the cross-sectional shape of the channels were broad and low rather than semi-circular (Hooke *et al*., 1990), or the ice were 'softer' than suggested, rates of channel closure would be greater than calculated here, and the extent of channels at the start of the melt season would be correspondingly reduced. Thus we conclude that subglacial channels would not survive the winter except beneath a short section of the glacier immediately up-glacier from the snout. They must therefore re-form each melt season, and this process must contribute to the rationalisation of the drainage system demonstrated by the dye tracing results.

Further support for this argument is provided by analyses of the chemistry of meltwaters draining from the glacier during the melt season. Meltwaters entering the glacier contain very little dissolved material, but their

solute load increases substantially as a result of contact with rock flour. Initial weathering of rock flour at Haut Glacier d'Arolla is by the carbonation reactions (Brown *et al*., 1994), resulting in Ca^{2+}/HCO_3^- waters. More prolonged water–rock contact allows sulphide oxidation to occur (Tranter *et al*., 1993). Some of the acidity generated by this reaction is neutralised by carbonate weathering, resulting in $Ca^{2+}/HCO_3^-/SO_4^{2-}$ waters. The ratio SO_4^{2-}/HCO_3^- can therefore be used as an index of mean rock:water contact time. Since meltwaters probably only come into contact with significant amounts of rock flour after they reach the glacier bed, this parameter can also serve as an indicator of mean meltwater residence time in the subglacial drainage system. A decrease in the ratio suggests a decrease in mean subglacial residence time. Between 11 June and 21 July, the ratio showed a steady decrease from 0·45 (standard deviation, 0·06) to 0·26 (s.d., 0·06) (Brown, 1991). These ratios are calculated means for 10 day periods based on twice daily sampling from the proglacial stream at the fluorometry station (Figure 1) at maximum and minimum discharge. These results suggest that at least part of the reduction in return time detected by the dye tracing experiments resulted from more rapid flow through the subglacial drainage system.

THE RATE AND GLACIER-WIDE PATTERN OF DRAINAGE SYSTEM EVOLUTION

As the melt season progressed, an increasingly efficient, channelised drainage system developed within and beneath the glacier (Figure 8a and b). If we assume that flow velocities of $>0{\cdot}5\,\mathrm{m\,s^{-1}}$ and dispersivities of $<10\,\mathrm{m}$ indicate fully channelised drainage, it is clear that the channelised system expanded headwards during the 1990 melt season until about 1 August. The dispersivity record suggests that the channelised system continued to extend after 1 August, but at a decreased rate. Irregularities in the form of the contours arise because of day-to-day variations in the magnitude of the meltwater discharge, and because moulins located a similar distance up-glacier may be well separated across the width of the glacier. Such moulins may be connected to different channels, and the rate of headward growth of different channels may vary.

Two methods were used to estimate the rate of channel extension.

(i) The $0{\cdot}5\,\mathrm{m\,s^{-1}}$ contour on Figure 8a is taken as an indicator of the position of the channel head. This suggests that the channel head lay about 1 km from the snout on 28 June, and that it extended rapidly up-glacier to reach 3·2 km from the snout by 31 July, after which its position stabilised. The mean rate of extension during July was *c*. $65\,\mathrm{m\,d^{-1}}$. The position of the channel head inferred using this approach correlates very well with the position of the transient snowline (Figure 8a). The significance of this, and the likely controls on the channel extension process, will be discussed below.

(ii) It is assumed that the flow path from injection site to snout is a composite one, comprising a section of distributed drainage system feeding into a channelised system. If each system has a 'characteristic' flow velocity, the length of flow path in the channelised system, (x_c) can be estimated from:

$$x_c = \frac{(x u_c - ((x/u_m) u_c u_d))}{(u_c - u_d)} \tag{3}$$

(Willis *et al*., 1990, equation 22). Here u_m is the mean flow velocity derived from a tracer experiment, u_d and u_c are the velocities in the distributed and channelised systems respectively, and x is the distance from injection site to snout.

Velocities in the distributed system between m8Ca and m7Cb in 1990 varied between 0·012 and $0{\cdot}068\,\mathrm{m\,s^{-1}}$ (Nienow, 1993). These values compare with velocities inferred to result from flow through inferred linked-cavity systems of $0{\cdot}025\,\mathrm{m\,s^{-1}}$ at Variegated Glacier during surge (Kamb, 1987) and $0{\cdot}03\,\mathrm{m\,s^{-1}}$ at Midtdalsbreen (Willis *et al*., 1990). A value of *c*. $0{\cdot}025\,\mathrm{m\,s^{-1}}$ therefore seems an appropriate estimate of u_d. Mean flow velocities within the channelised drainage system in the main catchment at the Haut Glacier d'Arolla are on the order of $0{\cdot}3$–$0{\cdot}5\,\mathrm{m\,s^{-1}}$, so values of u_c within this range were used. The exact value used was dependent on the mean discharge (Q_m) between the injection site and the glacier snout for each test. Although these values are poorly constrained, calculations using Equation 3 are relatively insensitive to variations in u_c because values of u_d are an order of magnitude lower.

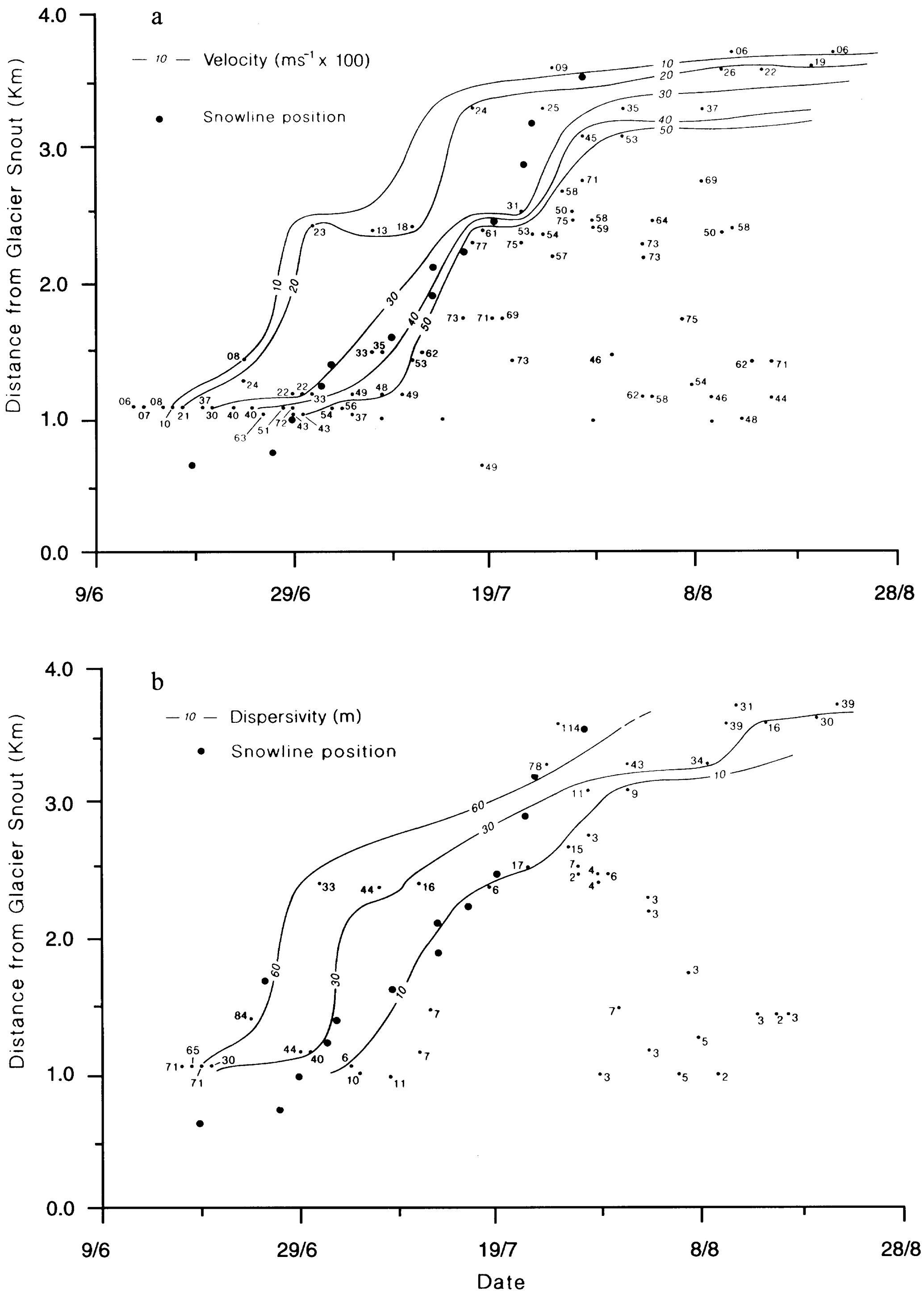

Figure 8. Contoured plots of (a) flow velocity, and (b) dispersivity as a function of location and date of injection for the 1990 melt season. Large dots indicate the position of the snowline at the glacier centreline

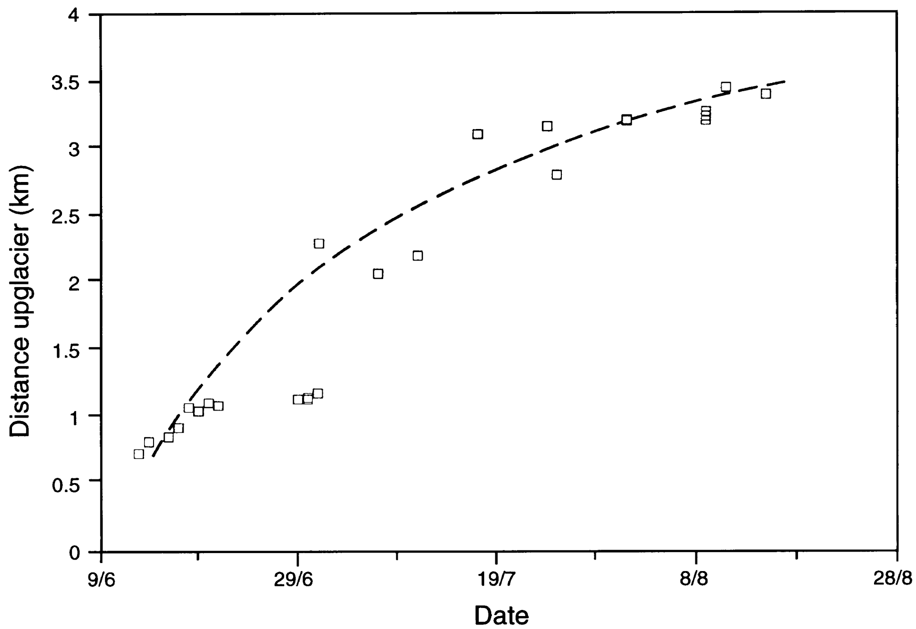

Figure 9. Position of the head of the channelised component of the drainage system during the 1990 melt season, as determined from observed dye return times and the assumption that water flowed through a two-component drainage system. Velocity in the distributed system = 0·025 m s^{-1}, velocity in the channel system is between 0·3 and 0·5 m s^{-1} depending upon meltwater discharge (see text for details)

Assuming $u_d = 0{\cdot}025$ m s^{-1}, application of this approach to the 1990 velocity results shown in Figure 3b suggests that, during the first test from each site, water spent between 64 and 94% of its overall travel distance in the channelised part of the system. The inferred position of the channel head moved from *c.* 700 m up-glacier from the snout on 13 June to a final position *c.* 3400 m from the snout (Figure 9). The rate of channel growth seems to have been about 60 m d^{-1} between 9 June and 19 July, decreasing to about 20 m d^{-1} thereafter. This suggests that the channel grew more rapidly in the lower glacier than in the upper glacier. If u_d is increased to 0·05 m s^{-1}, the inferred position of the channel head was just 200 m up-glacier from the snout on 13 June 1990, suggesting that there may have been only very limited channels beneath the Haut Glacier d'Arolla at the start of the melt season.

It is clear from this analysis of the growth of the channelised system that, at the end of the 1990 melt season, a residual distributed system survived up-glacier of moulin m7Ca (3·1 km up-glacier). In 1991, however, velocity results indicated that by 16 August, channelised flow conditions extended at least as far up-glacier as moulin m8Ca (3·6 km up-glacier). The extent of the residual distributed system therefore varied significantly between years.

PROCESSES AND MECHANISMS OF DRAINAGE SYSTEM EVOLUTION

The composite plots of flow velocity and dispersivity as a function of time and location of injection site in 1990 both suggest that there was a marked correlation between the approximate position of the channel head and that of the supraglacial snowline (Figure 8a and b). It thus appears that the retreat of the transient summer snowline determined the rate at which the channelised system expanded up-glacier. This suggests that the greater magnitude and diurnal amplitude of meltwater discharges associated with runoff from impermeable, low albedo ice surfaces (in comparison to discharges associated with runoff from permeable, high albedo snow surfaces) play a critical role in establishing major channels. This is consistent with the suggestion of Kamb (1987) that distributed linked-cavity systems may collapse into channelised systems if subjected to water pressure perturbations significantly above a steady-state value. Kamb (1987) envisaged distributed drainage

systems as consisting of a series of large cavities developed in the lee of bedrock obstacles, linked together by relatively narrow orifices. The linked-cavity system collapses due to the initiation of unstable orifice growth and consequent tunnel development resulting from significant water pressure perturbations above steady-state values (Kamb, 1987). Similarly, Walder and Fowler (1994) suggested that increasing discharge promotes the growth of incipient interfacial conduits into either R-channels incised into the glacier sole or canals incised into underlying sediment. The precise form of the final drainage system depends upon glacier surface slope, with R-channels being favoured at the high slopes characteristic of alpine glaciers.

The up-glacier retreat of the summer snowline could result in a mechanism capable of generating perturbations in water pressure sufficient to induce unstable orifice growth. Once the snowline has retreated up-glacier of a moulin, increased melt rates and rapid runoff from newly exposed ice surfaces would produce strongly peaked diurnal discharge cycles in the stream draining into the moulin. Discharges would be much greater than those entering the moulin whilst it was still snow covered (when melt rates would be lower and the storage capacity of the snowpack would smooth and delay the diurnal runoff hydrograph) and would be likely to induce a period of sustained high water pressures. This might be sufficient to initiate unstable orifice growth and channel formation.

An additional mechanism which might facilitate up-glacier channel growth is suggested by a consideration of the likely form of the piezometric surface in the vicinity of a channel head. If a major channel is able to form at the glacier bed, it is likely to transmit the available discharge at a lower pressure than the pre-existing distributed system (Walder, 1986; Kamb, 1987). In the area surrounding the channel head, there will therefore be a locally steep hydraulic gradient between the high pressure distributed system and the lower pressure channel system. This gradient is likely to result in the reorientation of water flow within the distributed system towards the channel head, and also in an increase in the local flow velocity within the distributed system. This increase in flow rate is likely to enhance the rate of ice melting in the vicinity of the channel head, and thus facilitate headward growth of the channel.

It is also likely that high discharges associated with early season rainstorms may initiate the formation of subglacial channels. Heavy rainfall on 16 and 17 June 1991 resulted in exceptional unseasonal runoff from the Haut Glacier d'Arolla drainage basin (peak discharge of $6{\cdot}1\,m^3\,s^{-1}$ at 22.00 hrs on 16 June; Figure 2b). The results of this storm were evident on the glacier surface, where two large channels were cut along the margins of the glacier tongue prior to drainage into moulins m3Eb and m3Ca (Figure 1). Channel growth induced by storm runoff may explain why the first dye injections carried out at moulins m3Ca and m3Eb in 1991 produced flow velocities typical of fully channelised flow conditions, even though the injection sites were located 0·5 km up-glacier of the snowline at the time.

Slowing of the rate of channel extension in the upper glacier (Figure 9) may be linked to the presence of a patchy firn aquifer on the glacier surface in this area. Such an aquifer might be expected to store meltwater, and to damp and delay the delivery of surface runoff to the englacial/supraglacial drainage system, even after the snowline has retreated further up-glacier. On a glacier with a thick continuous firn aquifer, this might be sufficient to halt the process of channel extension at the firn line. On Haut Glacier d'Arolla, however, a succession of negative mass balance years has eroded the firn aquifer, so that it is now patchy and poorly developed. Its impact on patterns of water delivery to the englacial/subglacial system appears to be sufficient to slow the process of channel extension, but not to halt it completely.

Two important points arise from the realisation that channel growth is linked to subglacial water pressure perturbations induced by changing patterns of supraglacial water input. Firstly, since the process of destabilisation of the distributed system is presumably initiated at the point where supraglacially derived waters first reach the glacier bed, the primary direction of channel growth is likely to be down-glacier. Therefore, the apparent pattern of headward channel growth that is documented here arises because up-glacier snowline retreat results in ice-derived meltwater inputs reaching the glacier bed at successively later dates further up-glacier. It is therefore likely that the process of channel growth creates continuous segments of channel extending up-glacier from the snout, which are separated by residual areas of distributed drainage from isolated segments of channel extending down-glacier from the moulin which has most recently started to receive ice-derived meltwater inputs. These segments extend progressively down-glacier until they link up with the continuous channel system. The process of channel growth will therefore be controlled by the distribution of

crevasses/moulins which enable precipitation and meltwater to enter the englacial/subglacial drainage system. Secondly, since both the distribution of moulins and crevasses across the glacier surface and the pattern of snowline retreat exposing such crevasses are irregular, it follows that the up-glacier expansion of the channelised drainage system is likely to be episodic in nature. Moreover, the rate and pattern of channel growth down-glacier from a given input point is likely to depend upon the magnitude and duration of the water pressure perturbation induced in the subglacial system. It is thus possible either that incipient channels will extend down-glacier from a crevasse to link up with the established channel system as a result of one major pressure event (such as the June 1991 storm event), or that down-glacier extension may occur over a period of several days and be driven by a series of smaller pressure events. In addition, as noted earlier, some channel growth may also occur in an up-glacier direction due to the effects of flow convergence around the channel head.

EVOLUTION OF THE CROSS-SECTIONAL AREA AND VOLUME OF THE DRAINAGE SYSTEM OVER A SUMMER MELT SEASON

Two key variables control the cross-sectional area of a drainage system: the discharge and the type of drainage system. With a given drainage configuration, the cross-sectional area would increase with discharge if the flow velocity remained constant. If the flow velocity increased with discharge, the cross-sectional area would increase more slowly, or even remain constant if the system was always water-filled. For a given discharge, the cross-sectional area necessary to convey the discharge increases as the flow velocity through the system decreases. The flow velocity is dependent on the drainage configuration and on the hydraulic gradient driving the flow. A conceptual model for the evolution of drainage cross-sectional area beneath a glacier must therefore take account of: (i) variations in discharge both along the length of the glacier and over time, and (ii) the changes in drainage system configuration which alter the velocity of water flow through the glacier.

On the basis of the evidence presented above, a model for the evolution of drainage system cross-sectional area in the main subglacial catchment basin at Haut Glacier d'Arolla was set up on the basis of the following assumptions: (a) the glacier is 4 km long and divided into eight segments each 0·5 km long; (b) the peak bulk discharge was 0·5 $m^3 s^{-1}$ on 14 June 1990, rising steadily to 5·0 $m^3 s^{-1}$ by 14 July and remaining constant thereafter. This pattern of discharge variation approximates to the 1990 record of peak daily discharges (Figure 2a); (c) the local discharge (Q_1) decreased up-glacier as described earlier. As the model was based on the main catchment basin, local discharge at the glacier snout was set to 80 per cent of the bulk value, decreasing to 0 per cent at the glacier head; (d) the subglacial drainage system contained channelised and distributed components, with respective flow velocities of 0·5 and 0·05 $m s^{-1}$; (e) up-glacier expansion of the channelised system occured at a rate of 65 m per day, with the channel head located 1 km from the glacier snout on 29 June (which, by extrapolation, places it 0·025 km from the snout on 14 June). The total cross-sectional area of the drainage system in the main catchment basin was then determined at 0·5 km intervals along the glacier every 5 days between 14 June and 28 August 1990 (Figure 10).

At any given location, the modelled cross-sectional area increased through time, because discharge increased and velocity remained constant, until either the distributed system converted to a channelised one or discharge became constant and the cross-sectional area of the distributed system stabilised. Since the channelised system expanded headwards, and had reached about 2·0 km from the snout when discharges became constant, the duration of the period during which the cross-sectional area of the distributed system remained constant increased up-glacier of 2·0 km. When the conversion from distributed to channelised drainage occurred, there was a sudden reduction in cross-sectional area. If the conversion occurred on or after 14 July, the cross-sectional area of channels remained constant (because discharge was constant after that date), but if it occured before this, the cross-sectional area increased as discharge rose over time. By 8 August, the drainage system had converted to a channelised form throughout the length of the glacier; this is an artefact of our failure to model the observed decrease in the rate of headward channel growth over time. After 8 August, the cross-sectional area at any given distance up-glacier remained constant because the local discharge was constant.

These results can be used to estimate temporal changes in the volume of the active part of the drainage system accessed by the tracer experiments by integrating the cross-sectional area of the drainage system over

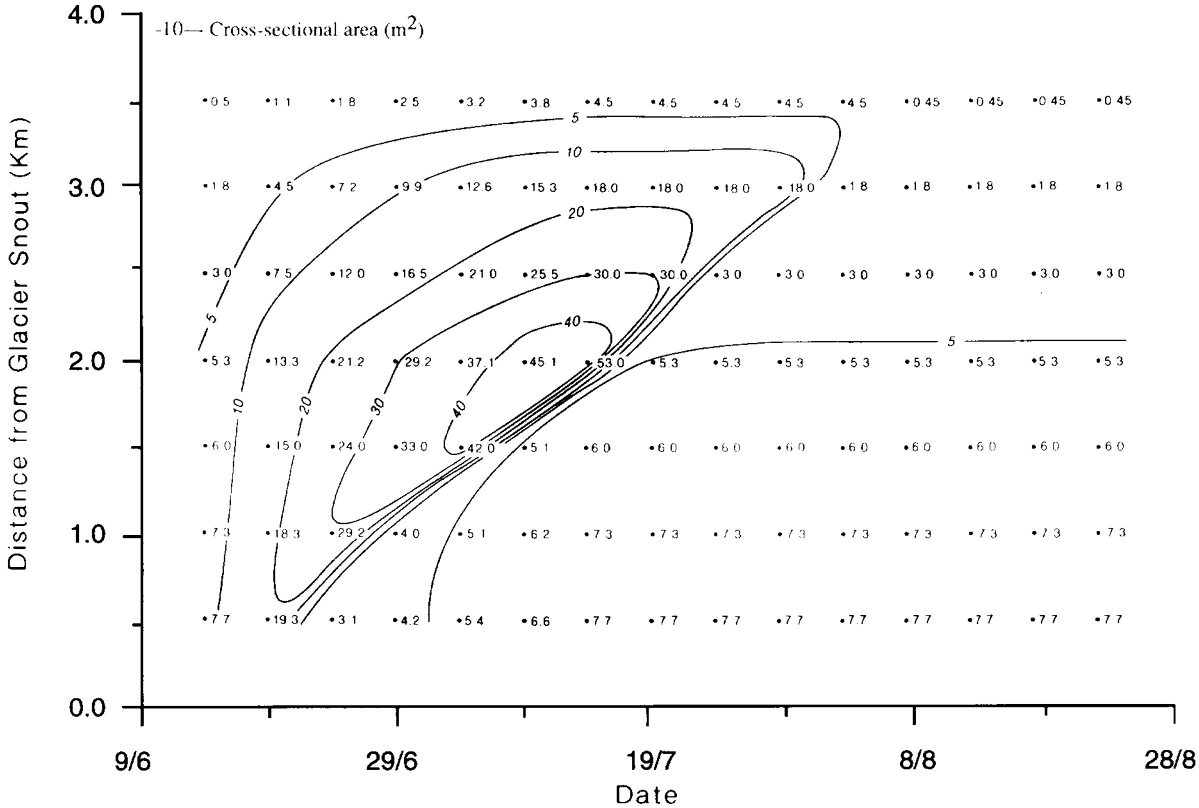

Figure 10. Model calculations of the distribution of drainage system cross-sectional area as a function of date and location beneath the Haut Glacier d'Arolla. Calculations are based on the 1990 melt season (see text for details)

the length of the glacier (cf. Smart, 1990). The volume increases from around 46 000 m^3 on 19 June, when the drainage system was primarily distributed in character, to 63 000 m^3 on 9 July when the channel head was located about 2 km from the snout. This suggests that during this part of the season, expansion of the distributed system in response to progressively rising discharges more than compensated for the reduction in system volume which resulted from channelisation of the drainage network. Between 9 and 29 July, the system volume decreased sharply to around 19 000 m^3 as a consequence of the eradication of the distributed system of the channelised system. This change in volume need not imply immediate collapse of the drainage pathways within the distributed system, but might simply suggest that many of these pathways ceased to be used by the bulk of meltwater discharge.

CONCLUSIONS AND DISCUSSION

Dye tracing results from the 1990 and 1991 melt seasons at Haut Glacier d'Arolla demonstrate significant glacier-wide evolution in the drainage system morphology. A system of major, hydraulically efficient channels seems to have expanded up-glacier over the course of each season at the expense of an inefficient distributed drainage system. Calculations suggest that channels developed during one melt season are unlikely to have survived to the next, except in a relatively small area extending no more than 200–700 m from the glacier snout. Elsewhere, ice deformation would have closed the channels before the start of the following melt season, leaving a residual distributed drainage system. At the onset of spring melt, a slowly varying input of meltwater from the supraglacial snowpack would have drained into this distributed system via moulins and crevasses. As the magnitude of this input increased, the cross-sectional area of the system would have increased to accommodate it, probably by a combination of ice melt, substrate erosion and cavity enlargement by enhanced basal sliding in response to rising subglacial water pressures. However, as the snowline retreated, exposing low

albedo, impermeable ice surfaces, moulins and crevasses would have received larger and increasingly peaked diurnal meltwater input cycles capable of generating high water pressure perturbations within the distributed system. These perturbations seem to have been sufficiently large or prolonged to destabilise the distributed system and concentrate drainage along a small number of pathways which evolved into major channels. Locally, channel growth was probably directed *down-glacier* from individual moulins. The overall direction of growth was, however, *up-glacier* because the retreating snowline continually exposed new moulins and crevasses and subjected them to diurnally peaked inputs of ice-derived meltwater. This continually shifted the locus of channel growth further up-glacier and allowed the channelised system to expand up-glacier at a rate similar to the rate of snowline retreat (up to 40–65 $m\,d^{-1}$). Since the snowline almost reached the glacier headwall in both seasons, the channelised drainage system dominated drainage of supraglacially generated meltwater acoss virtually the whole glacier by the end of each melt season. The rate of channel growth was, however, slowed by the presence of a discontinuous firn aquifer in the upper 1 km of the glacier, which damped down supraglacial meltwater inpluts and caused channel growth to lag behind snowline retreat.

These conclusions highlight three factors which are likely to control the evolution of subglacial drainage systems over summer melt seasons. These are the weather conditions during a particular season, the pattern of snowline retreat, and the distribution of moulins and crevasses which allow supraglacially derived waters to reach the glacier bed. The up-glacier extent of channel development is likely to be determined by a combination of how far up-glacier the snowline retreats, and how this relates to the distribution of the firn aquifer and of crevasses and moulins on the glacier surface.

The extent of snowline retreat will depend upon the distribution of supraglacial snow cover at the start of the melt season, and on weather conditions during the melt season and their impact on the rate of surface melt. In high-relief alpine terrain, the initial snow cover distribution is influenced by processes of wind drift and avalanching, whilst patterns of surface melt are strongly influenced by topographic and shading effects on the amount of solar radiation incident on the glacier surface. As a result, snowline retreat may be highly irregular and there may be differing degrees of drainage evolution beneath different parts of the same glacier. In some glaciers and some melt seasons, exceptional runoff events resulting from storm precipitation or jökulhläups may significantly affect the development of subglacial drainage channels. Under such extreme runoff conditions, a subglacial channel system may develop, the extent of which is totally unrelated to the position of the snowline.

The distribution of crevasses and moulins may exert a primary control on the degree to which the morphology of a glacier drainage system changes during the melt season. A heavily crevassed glacier with numerous entry points for supraglacial meltwater will probably deliver many low discharge inputs to the subglacial system. Even after exposure of ice on the glacier surface, these inputs might not be sufficiently large to produce the pressure perturbations necessary to destabilise a distributed drainage system, thus allowing it to persist throughout a melt season. On uncrevassed glaciers and those with a surface layer permanently below the melting point, surface melt may not reach the glacier bed. If the ice at the bed is at the pressure melting point, a drainage system immune to seasonal evolution may transmit water produced by basal melting to the glacier snout. On the other hand, dramatic changes in subglacial/englacial systems may occur in glaciers where large streams form on the glacier surface and enter the glacier at only a small number of input points. In such cases, the distribution of crevasses and moulins is likely to influence both the manner and rate at which the process of drainage system evolution occurs.

Another factor which may be important in determing the type of drainage system which develops beneath a glacier is the degree of similarity between the patterns of water flow on the glacier surface and the glacier bed. This will be a function of the subglacial bedrock topography, since the direction of subglacial water flow is determined primarily by the slope of the glacier surface except in cases where the slope of the bedrock surface is 10 times greater than that of the ice surface, and in the opposite direction (Shreve, 1972). At Haut Glacier d'Arolla, the form of the supraglacial drainage network and the anticipated pattern of water flow across the subglacial hydraulic potential surface are similar (Sharp *et al*. (1993). Thus supraglacial water inputs to the glacier bed are likely to be added directly to the main concentrations of water flowing from further up-glacier. This creates relatively large discharges which concentrate the generation of heat due to friction and loss of potential energy (and therefore melting) into restricted areas of the bed, and probably favours the formation of a

well-developed channel system. By contrast, in situations where there is no close correspondence between the form of the supraglacial channel network and the likely pattern of subglacial water flow, supraglacially derived waters may reach the bed away from preferred axes of subglacial drainage. The quantities of such waters may be insufficient to initiate channel formation where they reach the bed, causing them to drain significant distances through an inefficient distributed system before they combine with sufficient quantities of water following different drainage routes to initiate channel growth. This seems to be what happens at those moulins at the Haut Glacier d'Arolla which are not located close to major axes of subglacial drainage (Nienow *et al*., 1996b), and it may also explain some of the tracer results from Midtdalsbreen, Norway (Willis *et al*., 1990). It is particularly likely to occur in glaciers which have significant overdeepenings in their beds, since these cause divergence between the forms of the supra- and subglacial drainage systems.

ACKNOWLEDGEMENTS

This work was supported by NERC grant GR3/7004a, and by grants from Earthwatch. P.N. acknowledges receipt of NERC Studentship GT4/89/AAPS/53 and NERC Fellowship GT5/93/AAPS/1. We thank Dr W. H. Theakstone, University of Manchester, for the loan of a fluorometer, Grande Dixence S.A. for provision of discharge data and logistic support, and Yvonne Bams for all her help in Arolla. Field assistance was provided by Keith Richards, Chris Hill, Bryn Hubbard, Neil Arnold, Wendy Lawson, Julia Branson, Jim Strike, Stuart Lane, Jean-Louis Tison, Mark Skidmore, Nick Spedding and numerous Earthwatch volunteers and Cambridge undergraduates. In addition, we would like to thank David Sugden and two anonymous reviewers for their comments on an earlier draft of the manuscript.

REFERENCES

Behrens, H., Bergmann, H., Moser, H., Ambach, W. and Jochum, O. 1975. 'On the water channels of the internal drainage system of the Hintereisferner, Ötztal Alps, Austria', *Journal of Glaciology*, **14**, 375–382.

Brown, G. H. 1991. *Solute provenance and transport pathways in Alpine Glaciers*, PhD thesis, Southampton University, 1991.

Brown, G. H. and Tranter, M. 1990. *Hydrograph and chemograph separation of bulk meltwaters draining the Upper Arolla Glacier, Switzerland*, International Association of Hydrological Sciences Publication **193**, 429–437.

Brown, G., Sharp, M., Tranter, M., Gurnell, A. and Nienow, P. 1994. 'Impact of post-mixing chemical reactions on the major ion chemistry of bulk meltwaters draining the Haut Glacier d'Arolla, Valais, Switzerland', *Hydrological Processes,* **8**, 465–480.

Burkimsher, M. 1983. 'Investigations of glacier hydrological systems using dye tracer techniques: observations at Pasterzengletscher, Austria', *Journal of Glaciology,* **29**(103), 403–416.

Fischer, H. B. 1968. *Methods for predicting dispersion coefficients in natural streams, with applications to lower reaches of the Green and Duwamish Rivers, Washington,* US Geological Survey, Professional Paper, **582A**.

Fountain, A. G. 1993. 'Geometry and flow conditions of subglacial water at South Cascade Glacier, Washington State, U.S.A.; an analysis of tracer injections', *Journal of Glaciology,* **39**, 143–156.

Fountain, A. G. 1996. 'Effect of snow and firn hydrology on the physical and chemical characteristics of glacial runoff', *Hydrological Processes,* **10**, 509–522.

Hock, R. and Hooke, R. LeB. 1993. 'Evolution of the internal drainage system in the lower part of the ablation area of Storglaciären, Sweden', *Geological Society of America Bulletin,* **105**, 537–546.

Hooke, R. LeB, 1984. 'On the role of mechanical energy in maintaining subglacial water conduits at atmospheric pressure', *Journal of Glaciology,* **30**, 180–187.

Hooke, R. LeB. and Pohjola, V. A. 1994. 'Hydrology of a segment of a glacier situated in an overdeepening, Storglaciären, Sweden', *Journal of Glaciology,* **40**, 140–148.

Hooke, R. LeB., Laumann, T. and Kohler, J. 1990. 'Subglacial water pressures and the shape of subglacial conduits', *Journal of Glaciology,* **36**, 67–71.

Iken, A. 1981. 'The effect of the subglacial water pressure on the sliding velocity of a glacier in an idealized numerical model', *Journal of Glaciology,* **27**(97), 407–421.

Iken, A., Röthlisberger, H., Flotron, A. and Haeberli, W. 1983. 'The uplift of Unteraargletscher at the beginning of the melt season – a consequence of water storage at the bed?', *Journal of Glaciology,* **29**(101), 28–47.

Kamb, B. 1987. 'Glacier surge mechanism based on linked cavity configuration of the basal water conduit system', *Journal of Geophysical Research,* **92**(B9), 9083–9100.

Kamb, B., Raymond, C. F., Harrison, W. D., Engelhardt, H. F., Echelmeyer, K. A., Humphrey, N., Brugman, M. M. and Pfeffer, T. 1985. 'Glacier surge mechanism: 1982–1983 surge of Variegated Glacier, Alaska', *Science,* **227**, 469–479.

Nienow, P. W. 1993. *Dye tracer investigations of glacier hydrological systems*, PhD thesis, Cambridge University, 1993.

Nienow, P., Sharp, M. and Willis, I. 1996a. 'Velocity–discharge relationships derived from dye tracer experiments in glacial meltwaters: implications for subglacial flow conditions', *Hydrological Processes,* **10**, 1411–1426.

Nienow, P., Sharp, M. and Willis, I. 1996b. 'Temporal switching between englacial and subglacial drainage pathways: dye tracer evidence from the Haut Glacier d'Arolla, Switzerland', *Geografiska Annaler,* **78A**(1), 51–60.

Röthlisberger, H. and Lang, H. 1987. 'Glacial hydrology', in Gurnell, A. M. and Clark, M. J. (Eds), *Glacio-fluvial Sediment Transfer: an Alpine Perspective*, John Wiley and Sons, London, 207–284.

Seaberg, S. Z., Seaberg, J. Z., Hooke, R. LeB. and Wiberg, D. W. 1988. 'Character of the englacial and subglacial drainage system in the lower part of the ablation area of Storglaciären, Sweden, as revealed by dye-trace studies', *Journal of Glaciology,* **34**, 217–227.

Sharp, M. J., Richards, K., Willis, I., Arnold, N., Nienow, P., Lawson, W. and Tison, J -L. 1993. 'Geometry, bed topography and drainage system structure of the Haut Glacier d'Arolla, Switzerland', *Earth Surface Processes and Landforms,* **18**, 557–572.

Shreve, R. L. 1972. 'Movement of water in glaciers', *Journal of Glaciology,* **11**, 205–214.

Smart, C. C. 1990. 'Comments on: ''Character of the englacial and subglacial drainage system in the lower part of the ablation area of Storgaciären, Sweden, as revealed by dye-trace studies'', *Journal of Glaciology,* **36**, 126–128.

Spring, U. and Hutter, K. 1982. 'Conduit flow of a fluid through its solid phase and its application to intraglacial channel flow', *International Journal of Engineering Science,* **20**, 327–363.

Stone, D. B. and Clarke, G. K. C. 1996. '*In situ* measurements of basal water quality and pressure as an indicator of the character of subglacial drainage systems', *Hydrological Processes,* **10**, 615–628.

Tranter, M., Brown, G. H., Raiswell, R., Sharp, M. J. and Gurnell, A. M. 1993. 'A conceptual model of solute acquisition by Alpine glacial meltwaters', *Journal of Glaciology,* **39**, 573–581.

Walder, J. S. 1982. 'Stability of sheet flow of water beneath temperate glaciers and implications for glacier surging', *Journal of Glaciology,* **28**, 273–293.

Walder, J. S. 1986. 'Hydraulics of subglacial cavities', *Journal of Glaciology,* **32**, 439–445.

Walder, J. S. and Fowler, A. C. 1994. 'Channelised subglacial drainage over a deformable bed', *Journal of Glaciology,* **40**, 3–15.

Willis, I. C., Sharp, M. J. and Richards, K. S. 1990. 'Configuration of the drainage system of Midtdalsbreen, Norway, as indicated by dye-tracing experiments', *Journal of Glaciology,* **36**, 89–101.

Willis, I. C., Sharp, M. and Richards, K. S. 1993. 'Studies of the water balance of Midtdalsbreen, Hardangerjökulen, Norway. I. Water storage and runoff prediction', *Zeitschrift für Gletscherkunde und Glazialgeologie,* **27**(1991), 117–138.

